11. Icy Bodies, and the Physics of Ice

Icy and rocky bodies

The outer part of the solar system, beyond Mars, is dominated by the giant gas planets. However, they are accompanied by a plethora of small bodies - the asteroids, their satellites, Pluto and Charon - which are generally made of ices, rock or a mixture of the two. Thus in these regions the physics of ices and ice/rock mixtures is important. There are a great diversity of different types and features, but we can find similarities and groupings which tell us a lot about the processes involved in their formation and their physics and chemistry.

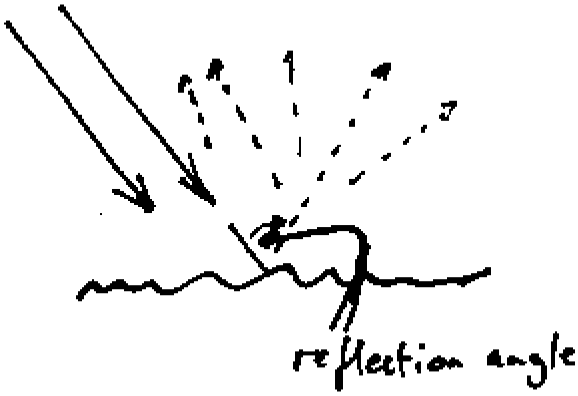
Remote sensing Techniques

To get information on these bodies we relied, prior to space probes to the outer system, on optical techniques. Even with space probes the chief method of gathering information is optical. We have all seen the magnificent photographs from the outer planetary systems of the larger (and occasionally smaller) satellites and rings, but other techniques are used besides the detailed photography, and these techniques are often useful in also telling us about the smaller bodies which cannot be photographed.

Getting Information from the Reflections from a body

Light curves

Light incident on a body's surface will be scattered, the scattering being dependent on the "roughness" of the surface features:



We can measure the intensity of reflected light for a body over a range of conditions - the so-called "light curve", which may be the variation of reflected light with time, angle to the surface, or phase angle at a given reflection angle (see below). For non-specular reflection we will get some reflection in all directions. If we integrate the reflected radiation over the whole sky for one frequency then the fraction returned is the monochromatic Bond Albedo A_{LAM} (which of course has to be less than 1). If we sum over all frequencies we can get the (overall) Bond Albedo A:

$$\sum_{\text{freq}} \frac{A_{\text{Lam}} \cdot \text{flux}}{\text{Total flux}} = \text{Bond Albedo } A$$

That A<1 is the response to solar heating - (1-A) must be re-emitted in the infra-red as the thermal flux from the body.

To measure the total flux from a body at all angles is not so easy in practice. If we define the phase angle, phi, as the angle formed by the path sun-satellite-observer, then we see we will be restricted for all bodies in orbit beyond Earth's to a small range of angles - the further out the more restricted. Thus the range of observable phase angles at Uranus is only +/- 3 degrees! Given that the observed flux at phase angle phi is F_{phi}, the phase integral is define the phase integral is defined as:

$$q = 2\int_{0}^{\pi} F_{\text{phi}} \sin(\text{phi})\, d(\text{phi})$$

A Lambert surface is a diffuse perfect reflector at all wavelengths. Then p, the geometric albedo is defined as:

$$p = \frac{\text{amount of scattered radiation}}{\text{amount of radn. from a Lambert body}}$$

where the actual and Lambert surface have the same cross-sectional area. Then A=pq.

A rough surface is one that multiply scatters light. A rough dark surface will have well-defined shadows due to trapping of light within the surface texture. For such a surface the brightness drops off rapidly away from the illumination angle. Thus the change of returned light intensity with phase angle will be different for different types of surface, and this behaviour can be used to categorise surfaces. Surface crystals will often have a different effect again. Light trapped in them may suffer multiple reflections before emerging, and this again will give varying responses with different angles.

Light curves of bodies can tell a lot about them indirectly. For example one can tell that most satellites are locked rotationally onto their primaries. As a satellite rotates one will usually get a variation of the reflected light, as lighter and darker areas are illuminated. This is often a complex pattern, but generally the pattern will recur every spin period. So by looking y spin period. So by looking at the recurrence time for the light curve we can tell the rotation period. For most satellites this is found to be equal to the orbital period. In the case of Titan there is very little variation, suggesting a dense, hazy atmosphere. Iapetus has a very peculiar behaviour, with 2 magnitudes of brightness variation as it rotates. The spectrum is found to be the same at all points of the rotatiom, but the albedo varies with longitude by a factor of 7!

Spectral information

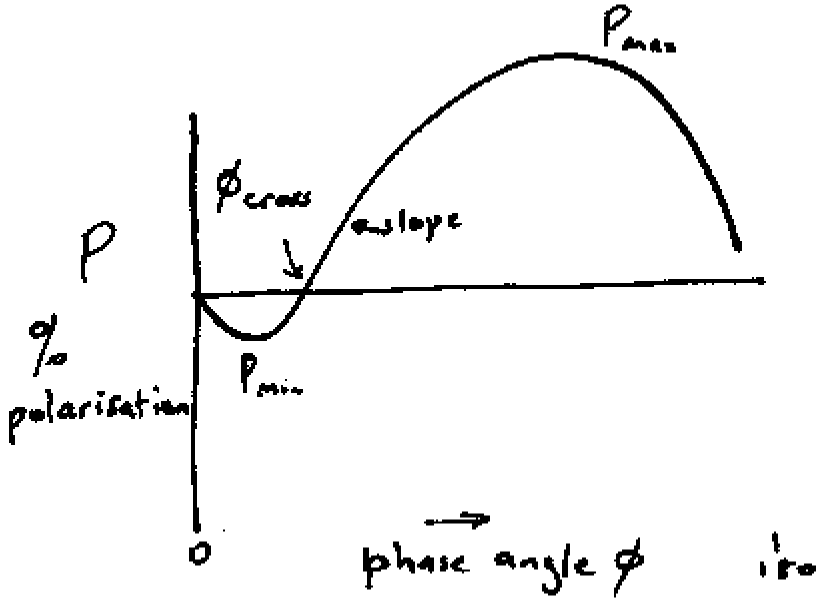
We can measure the visible and infra-red radiation separately with a radiometer. The visual brightness is proportional to A and the cross-sectional area, the infra-red brightness is proportional to (1-A) and the cross-sectional area. Thus by measuring both we can get A and the radius. Neglecting latitude effects the absorbed radiation must equal the emitted:

$$\underbrace{(1-A)F_{\odot}/R^2}_{\text{absorbed solar energy}} = 4\underset{\substack{\uparrow \\ \text{thermal emissivity} \\ (\sim 1.0)}}{\epsilon} T_s^4 = 4\sigma T_e^4$$

Most materials we will come across will have absorption bands in the visible or near infra-red, so the spectrum of the body (ie the power response as a function of wavelength) tells us something about the material of which it is made. Originally it was only possible to make observations in crude colour bands, so the astronomical "UVB" system of classification was used. Nowadays, however, it is possible to measure spectra with much more precision, virtually, often, to the extent of getting almost continous spting almost continous spectral response curves. With this information, it was possible even before satellite flybys of the Jovian satellites to say that Io, Amalthea and Europa were very red, and that Europa had strong water ice absorption bands, while Ganymede was darker in the visible and near infra-red than the others and also had strong water ice absorption features. Titan and Triton, on the other hand had no water ice features but did show strong Methane bands. Saturn's rings and the satellites of Uranus were dominated by ice.

Polarization curve

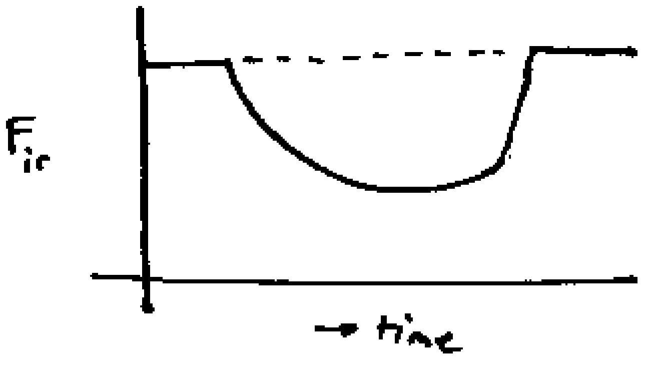
Another useful light curve is the polarization variation with phase angle:



The phase angle at which the curve crosses the x axis is usually 2-12 degrees depending on the surface. (Zero phase is equivalent to specular reflection.) The slope-albedo relationship is important especially to the study of asteroids. This technique has been also applied to the study of the Jovian satellites and to some of the Saturnian satellites through generally they are not bright enough. This technique has again suggested water ice on Ganymede, and for Io has suggested Sulphur, as well as SO_2 and H_2O. It has suggested water ice for Europa, and that Callisto is stony.

Eclipse radiometry

Eclipse radiometry uses the variation of thermal radiation with time as a body is eion with time as a body is eclipsed to tell something of the surface material. A typical flux/time curve for an eclipse transit looks as follows:



The rate of change of temperature with time gives some indication of the thermal inertia of the material from which it is made. For example, poor thermal inertia could be due to a poorly packed frost layer. The surface temperature variation plus some idea of the surface structure gives some idea of the thermal behaviour below the surface.

The thermal wave equation

If the heat flow through a volume changes with time the rate of change of temperature is given by the thermal wave equation:

$$\frac{\partial T}{\partial t} = -K\nabla^2 T = -K\left[\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2}\right]$$

where K is the coefficient of thermal conductivity. If we look at the penetration into a planetary surface of temperature variations due to an external heat source, then we can reduce this to one dimension:

$$\frac{\partial T}{\partial t} = -K\left(\frac{\partial^2 T}{\partial z^2}\right)$$

We shall assume the forcing function is periodic. It could be the diurnal variability of solar input, or the annual or seasonal variation of heating. We can also assume that the temperature varies with depth as f(z) (where z is depth). Then we can substitute for T(z,t):

$$T(z,t) = f(z)\, e^{i\omega t}$$

$$\text{so} \quad K\left(\frac{d^2 f(z)}{dz^2}\right) = i\omega f(z)$$

This has a general solution:

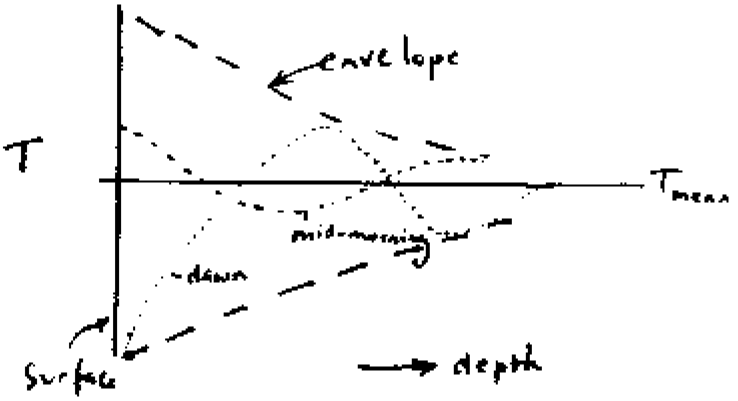
$$f(z) = Ae^{\alpha z} + Be^{-\alpha z}$$

The first term is physically meaningless as it represents a temperature exponentially growing with depth. Thus the equation for T(z,t) can be written:

$$T(z,t) = B e^{-\alpha z} e^{i\omega t}$$

or $$T = T_0 e^{-cz} \cos(\omega t - cz)$$

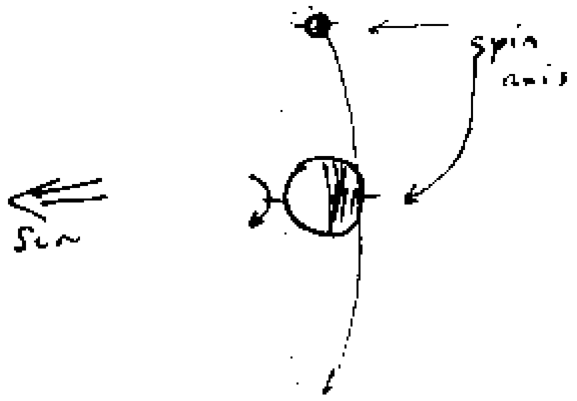
This is a sinusoidally travelling wave moving into the material from the surface, with an amplitude that drops exponentially with depth. Thus if we apply a sinusoidal variation at the surface, a little way below there will be a sinusoidal variation with a smaller amplitude and phase shifted from the input. The further in the larger the phase shift and the smaller the amplitude. The temperature at a great depth approaches the average value of the input variation. This equation can be used to explain things like the existence of permafrost. In the tundra regions, although the surface temperature may go well above zero in the summer months, a few meters below the surface it oscillates around the yearly average of the surface, and if that is below zero, the ice below the surface will never melt. Notice also that the peak temperature will be phase lagged compared to the surface - there will be a layer below the surface where the maximum temperature occurs at the point where the surface is at its coldest.



There will be different wavelengths and amplitudes for diurnal and seasonal waves. The exponential fall-off with depth is fairly fast with most materials - it is rare that waves get deeper than 10 m. In fact the diurnal wave usually only penetrates a few centimeters. Features of the structure of the topmost few meters may be influenced by the diurnal and annual thermal waves; even melting and evaporation may occur near to the surface, but the large-scale structure of the surface is rooted in "geological" scale processes deep in the interior.

The satellites of the outer planets: thermal trends

The daytime surface temperatures of the Galilean satellites are close to 100 K. That of the Saturnian satellites is about 70 K. The Uranian satellites are anomalous because of their geometry:



For half the Uranian year one axis or another of the satellites (as well as Uranus) is pointing at the Sun, so one side will be constantly heated and the other very cold. At the intermediate times, when the Uranian equatorial plane is edge on to the Sun, satellites and planet will get an evened out input over a rotation period. During Voyager 2's encounter the rotation poles pointed at the Sun, and temperature measurements of the sunlit face gave T_s elevated by square root of 2 times the "spherically distributed" heating situation. The dayside temperature pf Miranda was 86+/-1 K. The Planck peak of the nightside was so low in frequency it was not coven frequency it was not covered by the radiometer.

The surface temperature of Triton in the Neptunian system was the coldest so far seen in the solar system - 38 +/- 4 K. Smaller Neptunian satellites were not measured directly, but the albedos are lower so they are expected to be warmer.

Surface morphology of Satellites of the Outer Planets

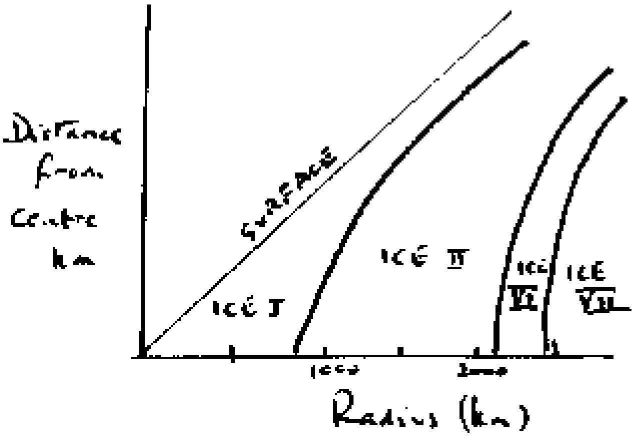
The Galilean satellites show several trends with distance from the primary. One trend is in the sort of surface morphology. Thus Io shows volcanism to the extent its surface is constantly being "remade", Europa shows some degree of activity, but some cratering survives on older surfaces, Ganymede has evidence on its surface of upwelling, but has a greater degree of cratering, and finally Callisto is heavily cratered and has littel sign of later surface activity (though it is smaller than Ganymede which may partly explain it.

The other trend is in density, which tends to fall off with distance from Jupiter, so that by Ganymede and Callisto there is more or less the solar proportions of ice and rock.

Saturnian satellites are all icy or ice-rock, with densities between 0.6 and 1.9 that of water. They are all heavily cratered. The Uranian satellites are also all low densities, with no evidence of a radial trend in densities. The Neprunian system is severely disturbed, and so is probably not "typical", but we can summarise the Jovian, Saturnian and Uranian ovian, Saturnian and Uranian systems by saying they are very "solar system"-like in their structures.

Physics and Chemistry of Ice

The physics and chemistry of ice is very important in the outer solar system. Ice has a number of forms, and the usual form of ice that we see on Earth - Ice I - is transformed under pressure to different crystal structures. Thus when a large body is built up from ice, or largely from ice, the type of ice of which it is composed will change with size. We can illustrate this with a diagram showing how increasing size produces increasing pressure and so an increasing tendency to produce Ice II, Ice III etc in larger and larger proportions:



This shows the self-compression effect under gravitational pressure, and assumes isothermal pure ice at 103 K. The lower the temperature and the more the impurities the lower the depth at which the transitions to higher forms take place, generally. Note that at around 800km and again at about 2000 km radius one expects discontinuities in the average density. This is true also of ice-rock mixtures at somewhat lower sizes.

The phase diagram for water is complex because of the different types of ice:

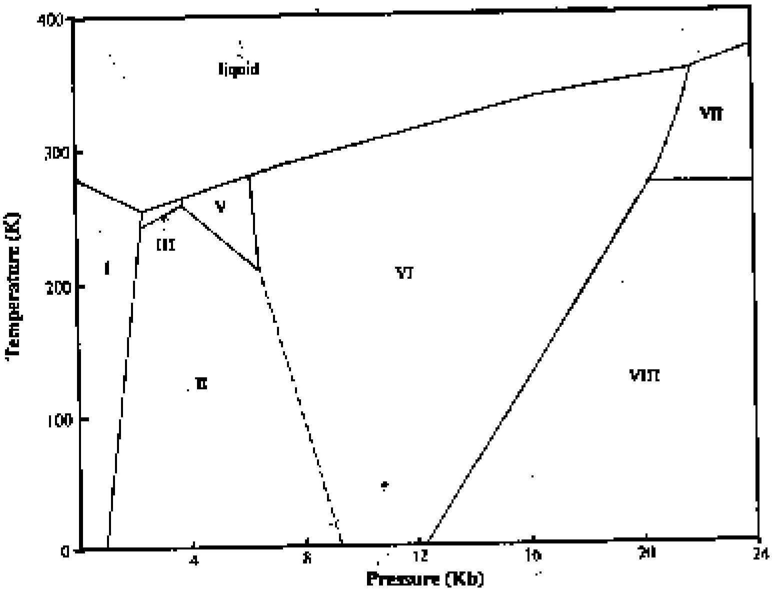


Figure VI.11 Phase diagram of water. The stability fields of liquid water, ordinary ice (I), and a number of high-density forms of ice are shown.

Phase diagram of water. The stability fields of liquid water, ordinary ice and a number of high density forms of ice are shown.

Water is "peculiar" in that ice I is less dense than the liquid - so down to -22 degrees C and 2115 bar pressure w degrees C and 2115 bar pressure will induce melting. At 'larger' pressures ice behaves 'normally' and would not be melted at any depth. At very high pressure you can get ice formed even at > 273 K.

Possible mechanisms for differentiation of the icy satellites

Because of this behaviour of ice it might be possible to get some differentiation of the icy satellites in regions where one might otherwise not expect the ice to be able to "flow". Of course, apart from pressure, there may be an alternative heat source in the form of radioactive elements inside the satellite. We can estimate how important this is likely to be, assuming solar-like composition of the original material. Then the initial rock and water ice planet/satellite will have nuclides providing about 1.6 10^{-11} J s^{-1} kg^{-1}, declining by a factor of 9 over the history of the solar system. The heat capacity of a rock/ice mixture is about 17 10^{3} J kg^{-1} K^{-1}. To heat this from 100K (typical Jovian system temperature) to 251 K (-22 C), the lowest eutectic melting point, requires $C_p$del-T (heat capacity times temperature change or energy required) of about 2.4 10^{5} J kg^{-1}, and so would take 5.8 10^{8} years, allowing for the fall-off in the decay rate, but assuming no heat was lost to the system to outside. Thus some significant heating might be possible, but the heat flux is always greater for solar flux than internal heat and only large bodies have chance to build up substantial temperatures. There is one helpful factor - that the thermal conductivity decreases as the temperature of ice increases - the temperature gradient becomes steeper with depth. However, the melting point decreases with pressure and at great depths you could not melt it.

Near the surface one might expect heat to be lost by radiation, so the key point where some liquefication, or at least fluidisation, might take place, might be expected to be some distance below the surface, but not too deep where melting is not possible. It turns out that the most interesting region in this respect is at a depth of about 80 km where the pressure is around 2280 bar, and the melting point is then at its lowest value, around 251 K. If the temperature of the surface is around 100K, then the temperature at 85km depth would be about 170 K from the nuclide heating in a typical model body. THis is not enough to melt the material at that depth, but it will be weakened and subject to viscous flow, which menas convective heat transport becomes possible.

Hence, if the satellite/planet islarge enough you could get a cold crust over a convective ice mantle over a rock core, even at Jovian distance from the Sun. The internal heat may even be increased by several mechanisms:

- early solar system radio-nucliar system radio-nuclide decay rate may have been higher
- there may have been other early heat sources (accretion/collapse etc)
- the freezing temperature can be lowered by impurities such as Ammonia
- there will be differing ratios of ice and rock
- there may be non-steady-state heat balance
- tidal energy can be a significant heat input especially for satellites near large bodies.

This last mechanism seems to be important for Io, where it probably provides the driving force for the satellite's volcanic activity, and maybe also Europa, where there is evidence that liquid - or at least viscous - water has flown at some period at or near the surface.

We can put the ideas above together to give a model for the differentiation of a large satellite of Jupiter (or maybe even Saturn with the right combination of circumstances). The diagram below shows the mechanism:this is taken from Lewis which should be consulted for a fuller description. The differentiation starts at that critical region about 85 km below the surface of the originally homogeneous body, but the convective transport helps the process of distributing the heat and continuing the differentiation process into the interior.

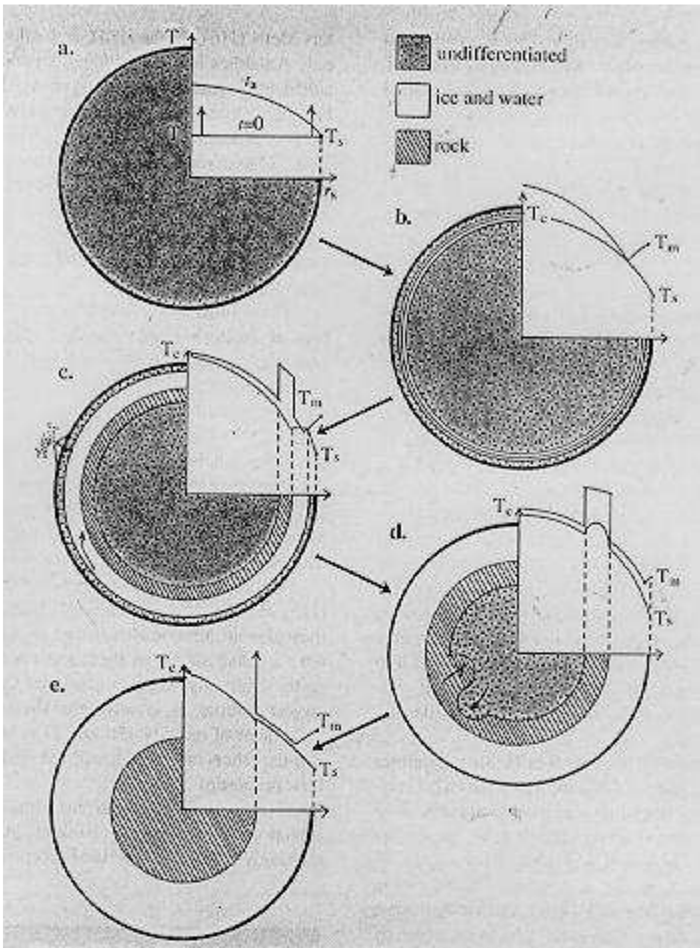


Figure VI.16 Differentiation of large ice-plus-rock bodies. (a) An initially homogeneous isothermal mixture of rock and water ice (here conservatively assumed to be at the surface temperature of the body) begins to heat up. (b) The internal temperature profile first reaches the melting curve near the 2-kbar pressure level, at which ice is easiest to melt. A thin layer segregates by density into dirt and water. (c) After more extensive melting, only an inner core and a thin crust remain undifferentiated. The crust, being denser than the pure water on which it floats, may founder as a result of faulting, impact puncturing, or volcanism. (d) The outer shell has now differentiated completely, leaving a thick outer layer of convecting pure water ice (with or without liquid water) atop a dense layer of rocky "mud," which in turn rests on a less-dense undifferentiated core. Convective instability inverts the interior structure, liberating energy that may complete the differentiation process (e). In general, the larger the body, the farther it can progress along this evolutionary path.

Fig VI.16 – Differentiation of large ice-plus-rock bodies.